# Crustal structure beneath the Trøndelag Platform and adjacent areas of the mid–Norwegian margin, as derived from wide-angle seismic and potential field data

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The outer mid-Norwegian margin is characterized by strong breakup magmatism and has been extensively surveyed. The crustal structure of the inner continental shelf, however, is less studied, and its relation to the onshore geology, Caledonian structuring, and breakup magmatism remains unclear. Two Ocean Bottom Seismometer profiles were acquired across the Trøndelag Platform in 2003, as part of the Euromargins program. Additional- land stations recorded the marine shots. The P-wave data were modeled by ray-tracing, supported by gravity modeling. Older multi-channel seismic data allowed for interpretation of stratigraphy down to the top of the Triassic. Crystalline basement velocity is ~6 km $s^{-1}$ onshore. Top basement is difficult to identify offshore, as velocities (5.3-5.7 km $s^{-1}$) intermediate between typical crystalline crust and Mesozoic sedimentary strata appear 50-80 km from the coast. This layer thickens towards the Klakk-Ytreholmen Fault Complex and predates Permian and later structur-ing. The velocities indicate sedimentary rocks, most likely Devonian. Onshore late- to post-Caledonian detachments have been proposed to extend offshore, based on the magnetic anomaly pattern. We do not find the expected correlation between upper basement velocity structure and detachments. However, there is a distinct, dome-shaped lower-crustal body with a velocity of 6.6-7.0 km $s^{-1}$. This is thickest under the Froan Basin, and the broad magnetic anomaly used to delineate the detachments correlates with this. The proposed offshore continuation of the detachments thus appears- unreliable. While we find indications of high density and velocity (~7.2 km $s^{-1}$) lower crust under the Rås Basin, similar to the proposed igneous underplating of the outer margin, this is poorly constrained near the end of our profiles. The gravity field indicates that this body may be continuous from the pre-breakup basement structures of the Utgard High to the Frøya High, suggesting that it could be an island arc or oceanic terrane- accreted during the Caledonian orogeny. Thus, we find no clear evidence of early Cenozoic igneous underplating of the inner part of the shelf.

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## Introduction

After the formation of the Caledonides, the Vøring Margin experienced several rifting episodes extending over a total of almost 300 m.y. culminating in continental breakup, followed by slight compression causing the formation of contractional domes (Fig. 1) (Blystad et al. 1995). Earlier rifting events were non-volcanic, but the final rifting and break-up was characterized by voluminous magmatism. The margin has attracted significant academic and industrial interest, due to the petroleum potential and the large variety of rifting processes that has contributed to the margin formation. Tectono-magmatic models for the crustal evolution of the northwestern part of the margin (Vøring Basin and Vøring Plateau) have been developed based on extensive coverage of industrial multi-channel seismic reflection (MCS) data, wide-angle seismic data acquired by the use of Ocean Bottom Seismometers (OBS), potential field data, and wells (e.g., Mutter et al. 1984; Eldholm et al. 1989; Skogseid et al. 1992; Osmundsen et al. 2002; Tsikalas et al. 2002; Gernigon et al. 2004; Mjelde et al. 2005a; Planke et al. 2005).

The sedimentary stratigraphy beneath the Trøndelag Platform and Halten Terrace is well established due to extensive coverage of MCS data and wells. However, our knowledge of the deep crustal structure along this part of the margin is limited. In order to fill this gap, two regional OBS-profiles were acquired across the Trøndelag Platform and Halten Terrace in 2003 (Fig. 1). The northern profile also crosses the Dønna Terrace. Both profiles terminate within the eastern Rås Basin. Additional seismic receivers were deployed onshore up to ~65 km from the coastline. The key elements these profiles were aimed at elucidating are the following:

- Depth to the crystalline basement and Moho (base of the crust): These represent key parameters in most basin models as well as in other types of tectonic or dynamic models.

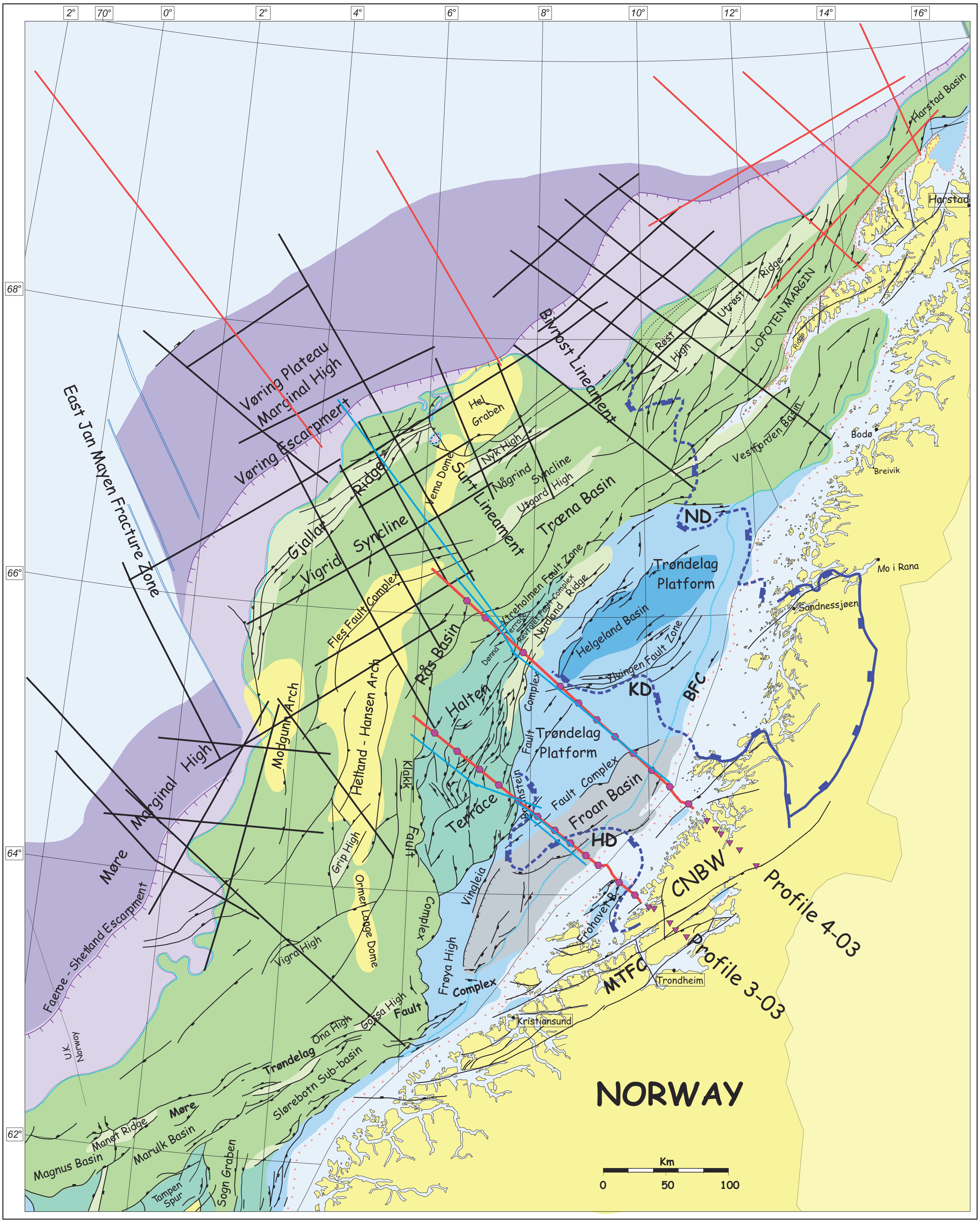


*Figure 1: The study area across the Trøndelag Platform, off mid-Norway. The shot lines of Profile 3-03 and 4-03 discussed in the paper are shown as bold red lines with purple hexagons marking OBS/OBH receiver positions, and inverted triangles marking land stations. Other OBS profiles from the Euromargins 2003 survey are shown by red lines, older surveys by black lines. The blue lines mark the location of individual MCS lines used for the composite stratigraphic section shown in Fig. 2. The structural base map is from Blystad et al. (1995), while the proposed offshore continuations of the Kollstraumen and Nesna detachments are from Olesen et al. (2002), and the Høybakken Detachment from Skilbrei et al. (2002). CNBW: Central Norway Basement Window, HD: Høybakken Detachment, KD: Kollstraumen Detachment, ND: Nesna Detachment.*

- Crustal composition: It is particularly important to reveal whether the high-velocity lower crust (~7.0-7.7 km $s^{-1}$) mapped in the Vøring Basin extends landward beneath the Trøndelag Platform or not. In the Vøring Basin this layer has generally been interpreted to represent mafic intrusions/underplating related to the last phase of rifting/break-up (e.g., Mjelde et al. 2005a). If igneous underplating occurs beneath the Trøndelag Platform, this would affect models for magmatism, extension, vertical movements and thermal evolution.

- Offshore extension of late- to post-Caledonian detachments: Our profiles are well suited to test a particular hypothesis regarding the offshore extension of some of these detachments, exposed onshore. The detachments were active during collapse of the Caledonian mountain belt and played a role in the denudation of the central Norway basement window (CNBW).

- To check whether the main Caledonian suture could be located beneath the Trøndelag Platform (e.g., Torsvik & Cocks 2005): Depending on the degree of overprinting by later extension, traces of the suture might be recognized by the experiment.

## Geological framework

Collapse of the Caledonides commenced during the waning stages of thrusting in the Early Devonian, (e.g., Andersen & Jamtveit 1990; Milnes et al. 1997; Fossen, 2000; Terry et al. 2000). The extensional detachment zones in Mid Norway demonstrate WSW-ENE bulk stretching axes, highly oblique to the mainly SE-directed thrust direction (Braathen et al. 2002). The Høybakken and Kollstraumen detachments frame the Central Norway Basement Window (CNBW), where highly metamorphosed Precambrian basement rocks border overlying Caledonian nappes (Braathen et al. 2000). Based on the interpretation of magnetic field data, it has been suggested that these detachments extend across the Trøndelag Platform (Fig. 1) (Olesen et al. 2002; Skilbrei et al. 2002). Though it is not stated clearly, the assumption must be that the drawn lines represent the intersection of the detachment plane with top basement, under the sedimentary succession. The two OBS profiles presented here are located between the Høybakken and Kollstraumen detachment traces, the southern profile crossing the proposed offshore Høybakken Detachment three times. Onshore, there are some minor Devonian deposits southwest of the Høybakken Detachment, and tongues of the Caledonian Uppermost Allochthon (Seve Nappes) are present between the detachments (e.g., Braathen et al. 2002; Osmundsen et al. 2003).

The margin developed through a series of rift phases before continental break-up in the Early Eocene (e.g., Eldholm et al. 1989; Brekke 2000). Important phases are: regional rifting during Late Carboniferous to Early Permian (~300-280 Ma), Triassic block faulting, and Late Triassic-Early Jurassic growth faulting (~220-200 Ma) (Surlyk et al. 1984; Blystad et al. 1995). Late Middle Jurassic to Late Cenomanian rifting created the deep Cretaceous basins of the outer margin (Brekke 2000). The Trøndelag Platform and Halten Terrace have deep pre-Middle Triassic half-grabens bounded by N and NE trending faults, detaching at upper- to intra-crustal levels (Osmundsen et al. 2002). These are likely related to the Devonian - Early Carboniferous post-orogenic extension of the Caledonian nappes, and subsequent reactivation of faults may have affected Permian to Early Triassic basin development. The Klakk Fault Complex and the Bremstein-Vingleia Fault Complex (BVFC; Fig. 1) are fundamental border faults in the Late Triassic-Jurassic rift system. (We will use FC for 'Fault Complex' and FZ for 'Fault Zone'). The BVFC marks the boundary between the Trøndelag Platform and the downfaulted Halten Terrace, whereas the Klakk FC represents the boundary between the Halten Terrace and the Rås Basin. The Trøndelag Platform comprises several sub-elements, of which our profiles cross the Froan Basin, bounded to the southeast by the Border FC. The southern profile also crosses the coast-near Frohavet Basin. Two wells drilled east of the Helgeland Basin north of our profiles documented a total of 750 m of Late Permian and Early Triassic sedimentary rocks (Bugge et al. 2002). The platform has been a stable tectonic area since the Late Jurassic rifting ended, and the relatively thin Middle Triassic - Cenozoic succession on the Trøndelag Platform consists largely of flat-lying or gently seaward-dipping strata (Blystad et al. 1995).

A minimum thickness of 3 km of Pre-Cretaceous sedimentary rocks is estimated regionally in the Vøring Basin (Bøen et al. 1984; Skogseid and Eldholm, 1989), while as much as 10 km may exist locally (Eldholm & Mutter, 1986; Planke et al. 1991; Raum et al. 2006). The Late Jurassic to Early Cretaceous extensional phases led to major faulting with reactivation of older fault zones (Bøen et al. 1984; Skogseid & Eldholm, 1989). Tectonic activity has also been invoked for the mid-Cretaceous evolution of the Norwegian Sea (e.g., Doré et al. 1999; Brekke 2000), although some authors argue in favor of Cretaceous tectonic quiescence until the Early Campanian rifting (Færseth & Lien, 2002).

The Early Cenozoic extension axis shifted westward with respect to the Late Jurassic-Early Cretaceous tectonic episode. The Late Cretaceous to Early Cenozoic extension lasted 20-25 m.y., until continental break-up occurred in the Early Eocene (54-55 Ma). Continental breakup was associated with voluminous igneous activity, generating both extrusions and intrusions into the adjacent sedimentary basin. There may have been massive emplacement of magmatic material at the base of the crust (Mutter et al. 1984; Skogseid et al. 1992; White & McKenzie 1989; Mjelde et al. 2005a), though the basis for this interpretation has been contested (e.g., Gernigon et al. 2003).

## Data acquisition and processing

The two OBS profiles were acquired in 2003, using R/V Håkon Mosby, University of Bergen. The survey was a cooperative venture between the Department of Earth Science, University of Bergen, the Department of Geosciences, University of Oslo, Institute of Seismology and Volcanology, Hokkaido University, Japan, and GEOMAR, Kiel, Germany. The source consisted of four equally sized Bolt airguns with a total volume of 78.66 l (4800 inch$^3$) fired at 200 m intervals. The 4.5 Hz OBSs deployed along the profiles were developed either at Hokkaido University (Shimamura 1996), or at GEOMAR (Bialas & Flueh 1999). Some of the GEOMAR instruments were equipped with hydrophones only (Ocean Bottom Hydrophones; OBH). A total of 12 instruments contained high-quality data along Profile 3-03 (285 km), whereas the acquisition along Profile 4-03 (356 km) provided high-quality recordings from 11 instruments. In addition, a total of 10 RefTek and Guralp 3C land stations owned by the University of Bergen were deployed along the landward continuation of each profile up to ~65 km south-eastwards. Due to timing problems on some of the instruments, only 5 good land-based data sets were obtained for each profile.

After linear clock drift correction, a 60 s record length for each shot was extracted and tied to the navigation. Navigation was performed by the Differential Global Positioning System. The instrument positions were relocated to correct for minor seacurrent induced drift along the profile direction. Initial processing consisted of de-biasing, band-pass filtering (6 to 12 Hz), and amplitude gain scaling (AGC). This was later compared to a processing flow consisting of de-biasing, predictive deconvolution, band-pass filtering, and either AGC or offset dependent scaling. Using a minimum lag of 80 ms in the deconvolution sometimes proved effective in damping the ringing from the bubble pulse while preserving weak arrivals. The P-wave data time-axis is velocity reduced by 8.0 km s$^{-1}$ and the data are AGC scaled for all examples.

## Velocity modeling

The wide-angle seismic data obtained from the OBSs do not image the underground like MCS data. Many of the observed arrivals are refracted through the crust at various levels rather than reflected. In order to make a geologically interpretable crustal transect, a model of the velocity distribution throughout the crust that can reproduce the observed seismic travel times at different offsets from the instruments, must be made. The model must satisfy all instruments in this respect, and having a number of them along the profiles helps to constrain possible models since the seismic energy travels through the same areas in the underground at different angles.

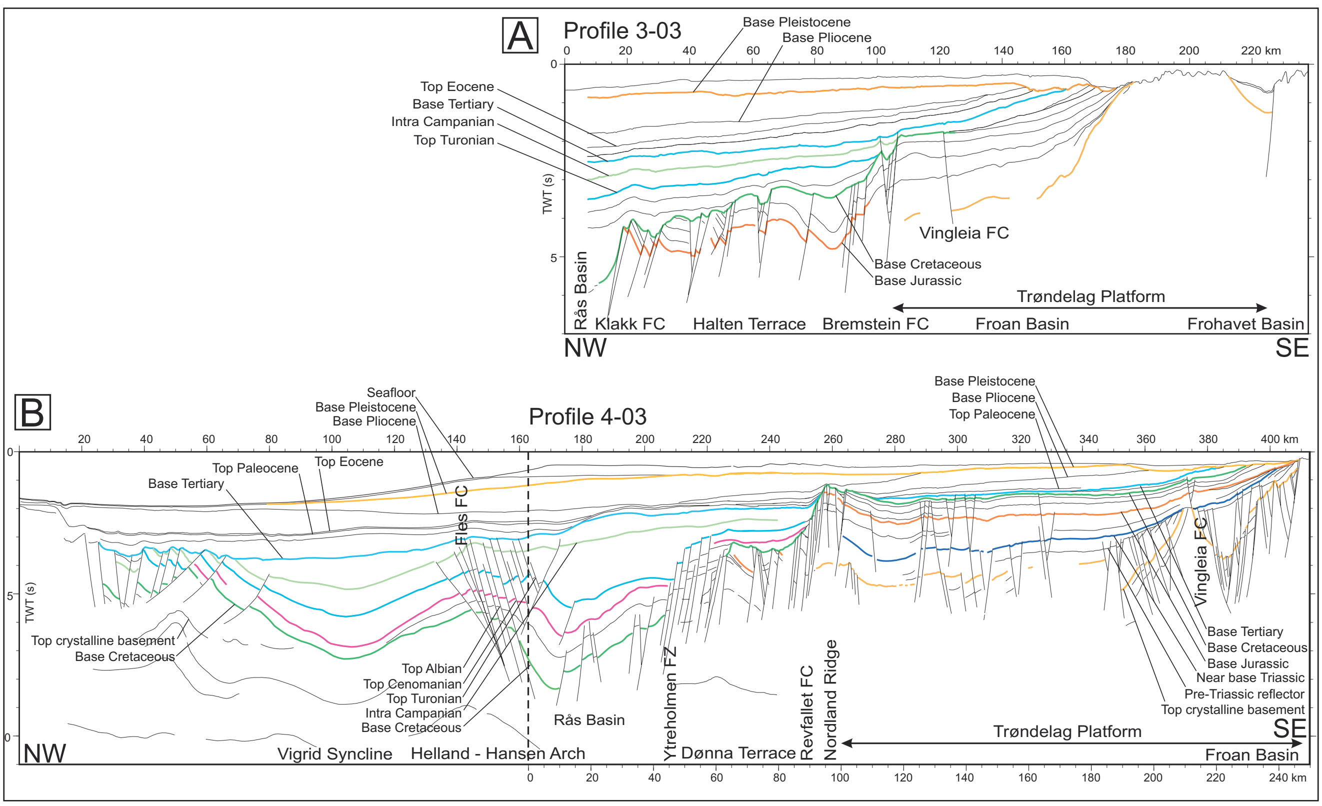


*Figure 2. Stratigraphic line interpretation of the MCS lines along the OBS profiles. A) Interpretation along Profile 3-03. B) Interpretation along Profile 4-03, with the OBS model distance scale shown at the bottom.*

The OBS data were acquired along MCS profiles collected earlier (Figs. 1 and 2). Key horizons were tied to a regional grid of MCS lines, and to neighbouring wells which assure reliable dating down to top Triassic on the Trøndelag Platform. The interpretation of deeper reflectors is more uncertain. The MCS interpretation was depth-converted using velocity information from wells, stacking velocities, and older Expanding Spread Profiles in the area.

These depth converted interpretations were used as initial models for the upper part of the OBS models. P-wave reflection and critical refraction travel-times were picked from the OBS vertical components, and modeled by combined forward (ray-tracing) and inversion (Zelt & Smith 1992). The modeling was done downwards, layer by layer. Stratigraphy does not necessarily follow velocity layering, and the initial model was simplified and adjusted to some degree. Thus, the final model layering does not represent an exact stratigraphic model. In general, there is a good correlation between burial history and velocity, due to the effect of compaction and diagenesis on seismic velocity (e.g., Domenico 1984). The crystalline basement was divided into four layers for both models so as to acieve the best fit for the data. Reflections that were not easily tied to layers were modeled by floating reflectors.

The MCS data give a good indication of the sedimentary rock distribution in the upper part of the model. Top basement and the crystalline crust are constrained by the OBS data. There is, however, variable resolution of different parts of the model due to differences in ray coverage and data quality. As there is a good correlation between seismic velocity and density (e.g., Ludwig et al. 1970) gravity modeling was used interactively to resolve ambiguities in the seismic data, and to constrain areas with poor seismic coverage. We will first discuss the seismic modeling of each profile in detail below, starting with Profile 4-03 in the north. All data and models are documented in the supplementary material accompanying this paper.

*Profile 4-03*

There is good stratigraphic control down to near top Triassic on the Trøndelag Platform, while the depth to base Cretaceous is uncertain in the Rås Basin (e.g., Brekke 2000). The MCS section along the profile shows the Froan Basin as a distinct sub-basin, bordered to the west by a basement ridge at the Vingleia FC (Fig. 2B). The stratigraphy of the rest of the platform is flat-lying until it climbs gently into the Nordland Ridge, which is bordered to the west by the Revfallet FC. Farther west, the Dønna Terrace is bordered by the Ytreholmen FC towards the deep Rås Basin. We show data examples from OBSs 46 and 37 (Figs. 3 and 4), land station L-9 (Fig. 5), and the resulting velocity model in Fig. 6.

Onshore basement velocity is typically 6.0-6.1 km $s^{-1}$. The modeling indicates the appearance of a thin layer with intermediate velocity (~5.1-5.3 km s-1 $s^{-1}$) 10 km from the coastline, overlying basement. The layer clearly predates the formation of the Froan Basin. Velocities become somewhat higher (5.45 to 5.8 km $s^{-1}$) west of the Vingleia FC, and the layer forms the core of the Nordland Ridge. The velocity is greater than the velocity of the Cretaceous sedimentary units found at similar depths (4-7.5 km) in the Rås Basin, which is typically 4-4.5 km $s^{-1}$. Therefore this sequence probably represents well-consolidated sedimentary strata which could, at least in part, date from the Devonian. Velocities are similar to those of Devonian strata in East Greenland (Schlindwein & Jokat 1999, 2000). The layer may be ~2-3 km thick at the Trøndelag Platform, but the transition to higher velocity below is not very distinct. Velocities below reach 5.9-6.1 km $s^{-1}$ at ~8 km depth under the central Trøndelag Platform, indicating basement depth.

The OBS data show some reflected arrivals from within the deeper parts of the Rås Basin on OBSs 45 and 46. These indicate a basement depth of ~11 km. This interpretation is supported by an upper basement arrival with high velocity on OBS 46 (Fig. 3, $P_{g2}$ on the left side). The MCS data show a pre-Cenozoic ridge with folding and uplift of strata into the Helland-Hansen Arch (Fig. 2B). A fit can be obtained by using a seismic velocity of ~7.2 km $s^{-1}$ combined with a rising basement here. The gravity modeling discussed below indicates that there is a substantial high-density body within the lower crust, but the seismic data do not cover the crust here and they are only able to identify this body within the upper part of the basement (Fig. 6).

Of the two faults between the Trøndelag Platform and the Rås Basin, the Revfallet FC at the Nordland Ridge has the strongest signature in seismic arrival times. An example of this can be seen in Fig. 3 at ~100 km, indicating that the fault complex involves more basement displace-ment than the Ytreholmen FZ between the Dønna Terrace and the Rås Basin. The velocity model indicates that displacement can be 4-5 km on the inner fault, and ~2 km on the outer fault (Fig. 6).

Crustal thickness is constrained by a combination of reflections from the Moho ($P_MP$) and refractions through the uppermost mantle ($P_n$). A $P_n$ phase arriving between 20 and 50 km on OBS 37 constrains the Moho depth under the Rås Basin to ~18 km (Fig. 4). OBS 37 also has a Moho reflection near the middle of the profile, which combined with $P_MP$ and $P_n$ phases from other stations (e.g., Fig. 5), constrains the Moho depth under the Trøndelag Platform to ~23-25 km. OBS 37 shows a lower-crustal reflection ($P_GP$) not tied to the velocity layering, consistent with reflections on OBSs 34 and 40. Shear zones and small-scale heterogeneities can cause reflectivity in the lower crust (e.g., Hurich & Smithson 1987). On the southeastern branch of OBS 37 (Fig. 4),

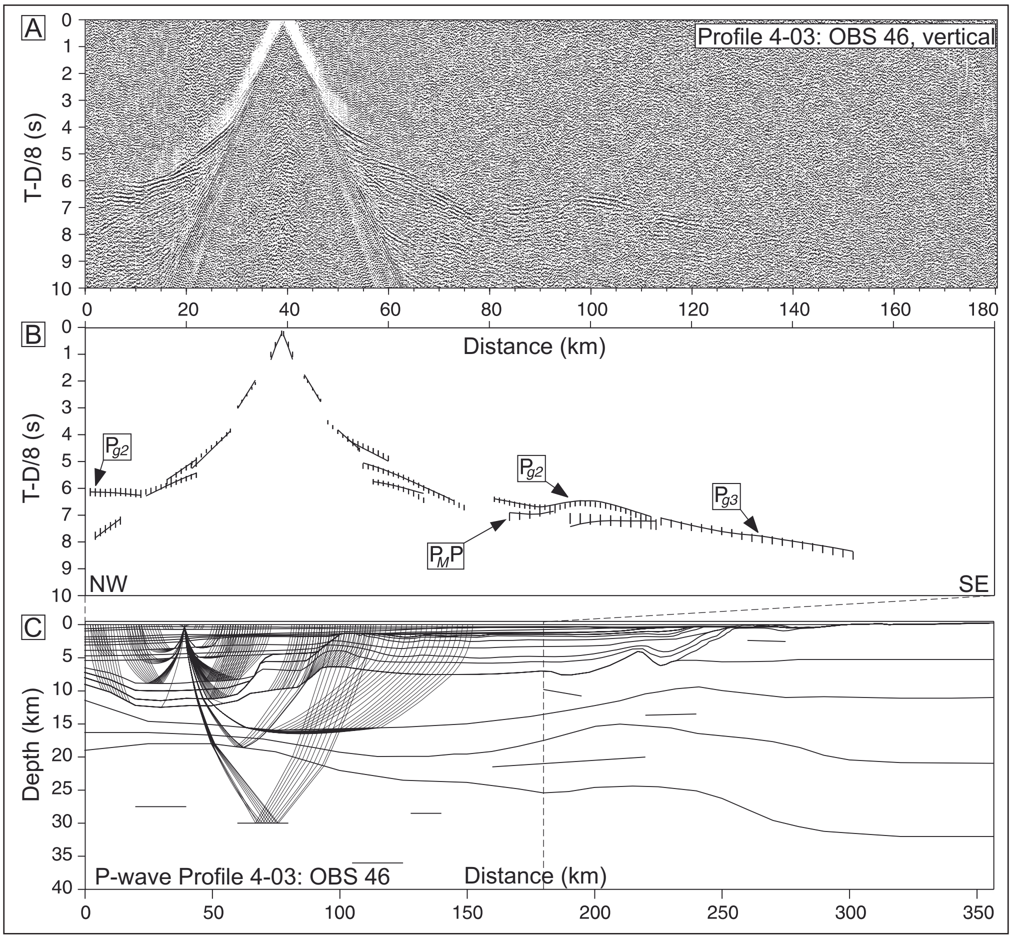


*Figure 3. Data, interpretation, and ray tracing of OBS 46, Profile 4-03.*
*A: OBS data, vertical component processed with spiking predictive deconvolution.*
*B: Interpretation and model reproduction.*
*C: Ray-tracing of the -velocity model.*

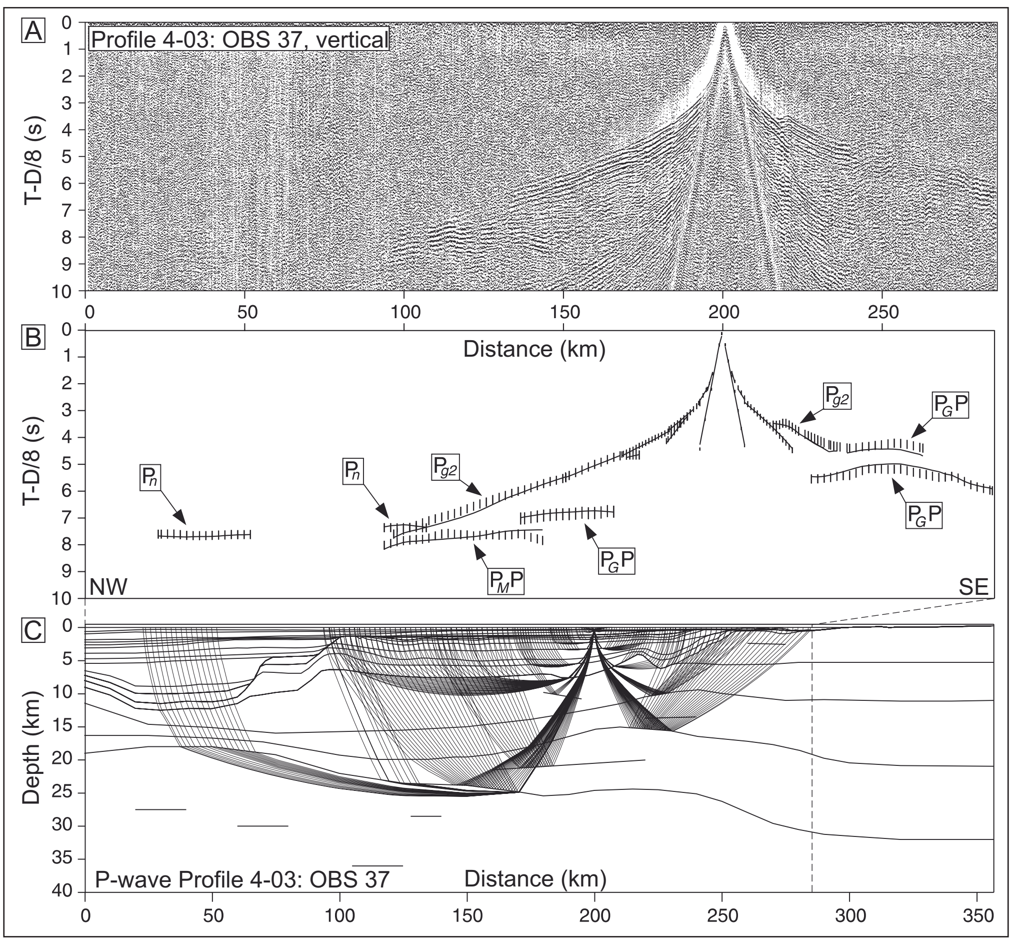


*Figure 4. Data, interpretation, and ray tracing of OBS 37, Profile 4-03.*
*A: OBS data, vertical component- processed with spiking predictive deconvolution.*
*B: Interpretation and model reproduction.*
*C: Ray-tracing of the velocity- model.*

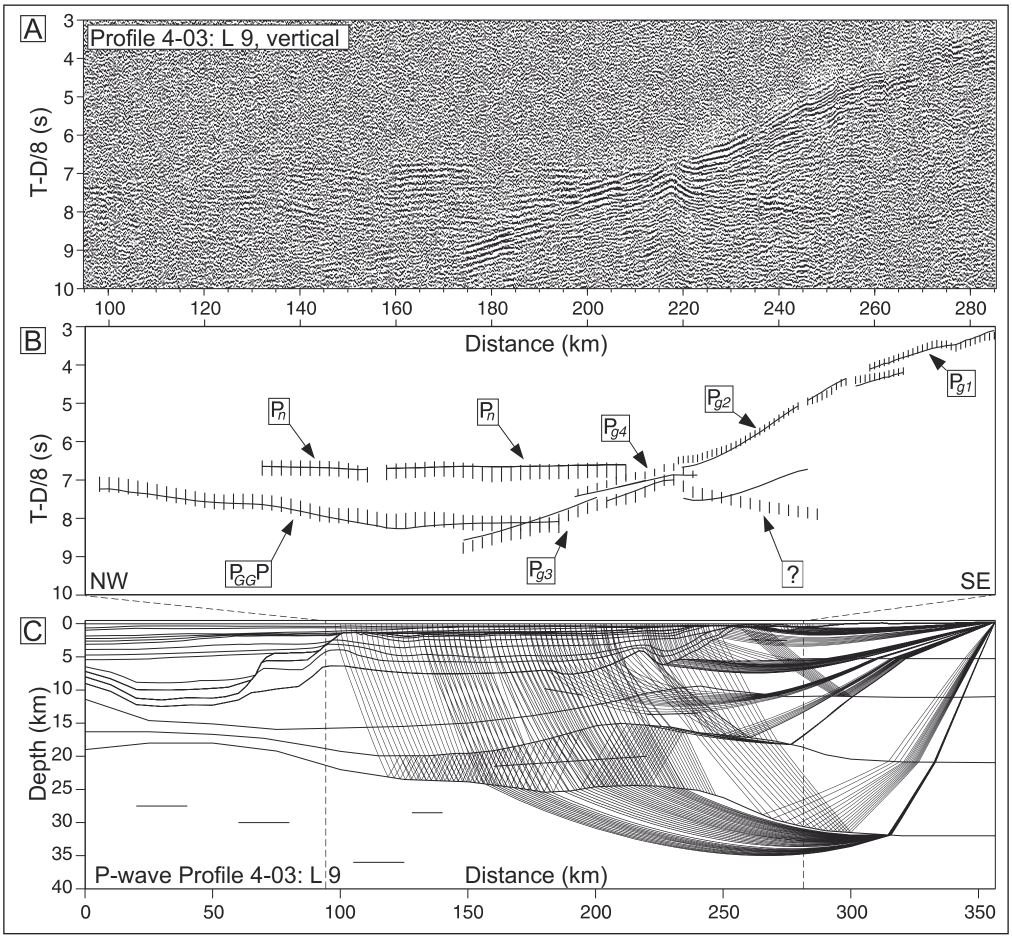


*Figure 5. Data, interpretation, and ray tracing of land station 9, Profile 4-03.*
*A: Data, vertical component processed with spiking predictive deconvolution.*
*B: Interpretation and model reproduction.*
*C: Ray-tracing of the velocity model.*

there is a high-amplitude reflection from the top of the lowermost basement layer. Several instruments show strong reflectivity from the top of this layer, as well as increased velocity. Land-station L-9 shows a high-amplitude $P_MP$ phase and a good $P_n$ phase, indicating a crustal thickness of 31-32 km beneath the coast. This instrument also shows a long offset phase with apparent high velocity after the $P_n$ phase ($P_{GG}P$, Fig. 5). This could have two different solutions. The first tried was similar to the model proposed by Christiansson et al. (2000) off western Norway. By introducing an eclogite layer within the lower crust, combined with a deeper Moho with a stronger dip towards the Trøndelag Platform, both phases could be reproduced. This complex model was, however, not compatible with the gravity modeling. The solution we believe to be the correct one is to model the arrival as a peg-leg multiple between the Moho and the top of the distinct lower-crustal layer. The observed arrival times are then well reproduced (RMS $\Delta T$ of 104 ms). We note that both the top and the bottom (Moho) of the body are good reflectors for direct arrivals (e.g., Fig. 4), and could thus also produce detectable multiples.

The Moho depth is 31-32 km at the coast, rising to 23-25 km underneath the Trøndelag Platform. There is a local 1-1.5 km increase in depth under the platform, likely related to the thinning of the crust underneath the Froan Basin to the southeast. The Moho rises farther to a depth of ~18 km under the Rås Basin, and the upper mantle velocity is normal (8.0 km $s^{-1}$).

*Profile 3-03*

Profile 3-03 is located ~100 km south of Profile 4-03 (Figs. 1 and 2A). The profile starts in the northwest just outside the Klakk FC, and covers the central parts of the Halten Terrace. From there it crosses the Bremstein FC and central parts of the Froan Basin. Mesozoic sedimentary strata onlap basement on the eastern side of the Froan Basin. The small, coast-near, Frohavet Basin (a half-graben) lies farther to the southeast (Blystad et al. 1995).

The velocity of the sedimentary section follows burial depth more than stratigraphy, though there is lower velocity with increasing depth basinward (Fig. 7). A velocity inversion below the Pliocene seen in wells farther- north (Reemst et al. 1996) shows as offsets in arrival times of refractions at 8-10 km distance (e.g., OBS 30). It is less clear on Profile 4-03 data, but visible on OBS 46 (Fig. 3). No structuring tied to the Froan Basin can be seen here. There is little velocity change across the Bremstein FC, and the throw may be as low as 1-2 km. The Klakk FC has a stronger traveltime signature, and the throw is estimated to ~5 km. The basement depth in

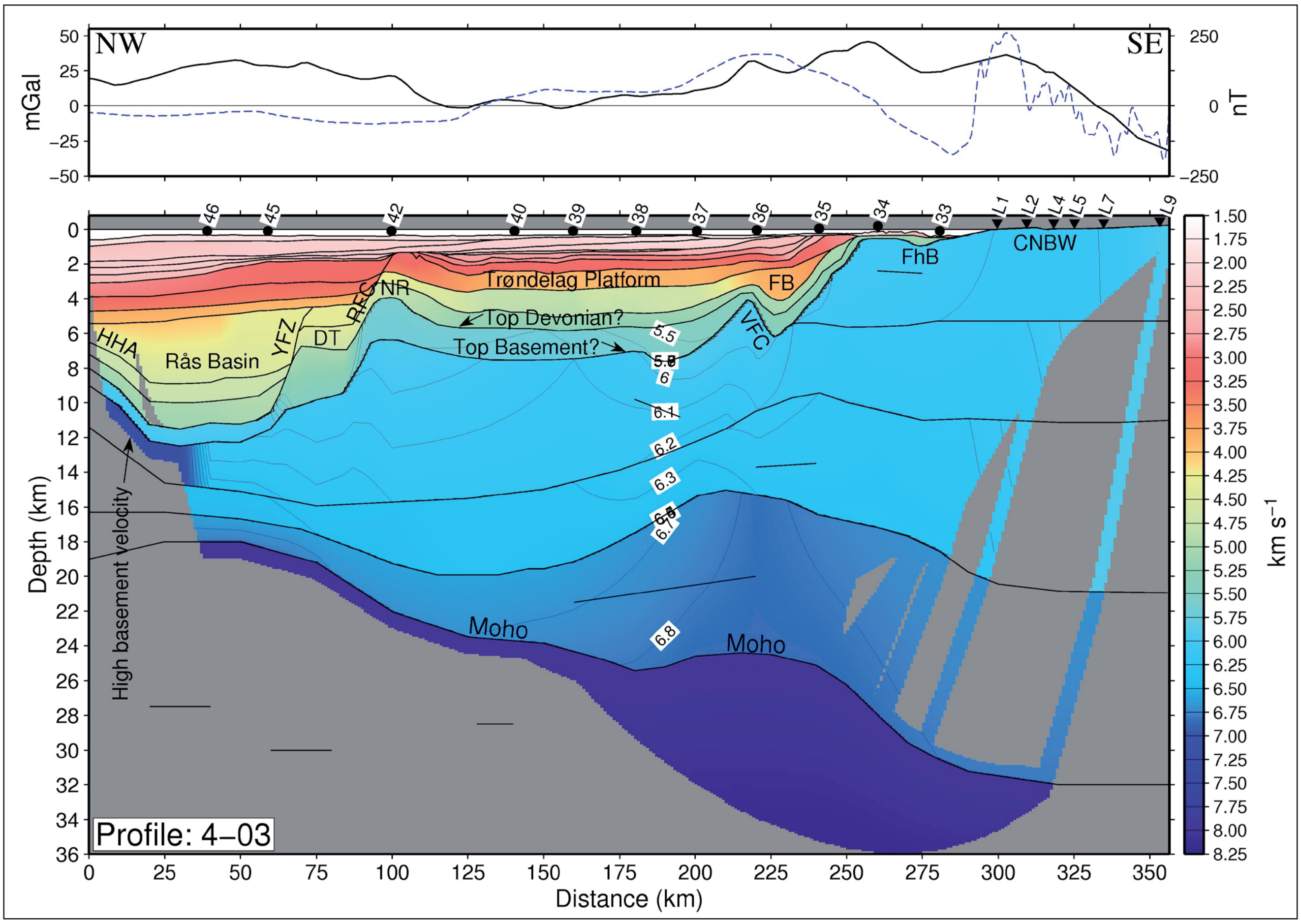


*Figure 6. Upper panel: Observed gravity (heavy line) and magnetic anomalies (dashed blue line) along Profile 4-03. Lower panel: Gridded crustal velocity model of Profile 4-03. The parts of the model not covered by rays are masked. Floating reflectors are not included in the ray-coverage. Contour interval within the basement is 0.1 km $s^{-1}$. CNBW: Central Norway Basement Window, DT: Dønna Terrace, FB: Froan Basin, FhB: Frohavet Basin, HHA: Helland-Hansen Arch, RFC: Revfallet Fault Complex, VFC: Vingleia Fault Complex, YFZ: Ytreholmen Fault Zone.*

the Rås Basin is poorly constrained: Arrivals traveling through the basin are significantly delayed, indicating at least 11 km depth, comparable to Profile 4-03.

The upper basement velocity onshore is highest at the coast (6.05-6.10 km $s^{-1}$), marking the top of a distinct zone with higher velocities than the surrounding rocks, which possibly extends down to 8-10 km depth. At ~160 km in the model there is a rapid lateral transition from 5.95 km $s^{-1}$ to a layer with intermediate velocity (5.4-5.8 km $s^{-1}$) farther west. This layer predates late Paleozoic - Mesozoic structuring, and obscures top crystalline basement on the platform. As for the similar layer on Profile 4-03, this is interpreted as well-consolidated predominantly Devonian sedimentary rocks. The depth to this layer is 6-7 km at the Halten Terrace, with a local rise adjacent to the Klakk FC, and about 5 km at the Trøndelag Platform. There is no clear transition to basement below this layer, and we do not try to quantify basement depth on the platform for this profile.

Profile 3-03 has a distinct lower crust with velocity of 6.5-7.0 km $s^{-1}$ in central parts of the model, similar to that on Profile 4-03. The Moho depth is well determined between 20 and 180 km in the model. It rises evenly from 31 km depth at the coastline, to 22 km at the western edge of the Halten Terrace, but both Moho depth and basement are unconstrained under the Rås Basin. Upper mantle velocity is 7.9 km $s^{-1}$ in the west and 8.1 km $s^{-1}$ at the middle of the profile, which is within the normal range (e.g., Holbrook et al. 1992; Mjelde et al. 1997).

*Uncertainties and resolution*

The uncertainties in seismic modeling can be represented by the $\chi^2$-value, which weighs the mismatch between observed and calculated travel-times to the estimated pick uncertainty so that $\chi^2 \leq 1$ signifies a fit (Zelt & Smith 1992). The assignment of uncertainty is somewhat subjective, including a number of issues where the ease of picking first onset is important (e.g., Hooft et al. 2000; Breivik et al. 2009). Arrivals from the upper-crust are assigned ±50-100 ms, while arrivals from mid-crustal levels are given ±100-200 ms, and Moho arrivals are

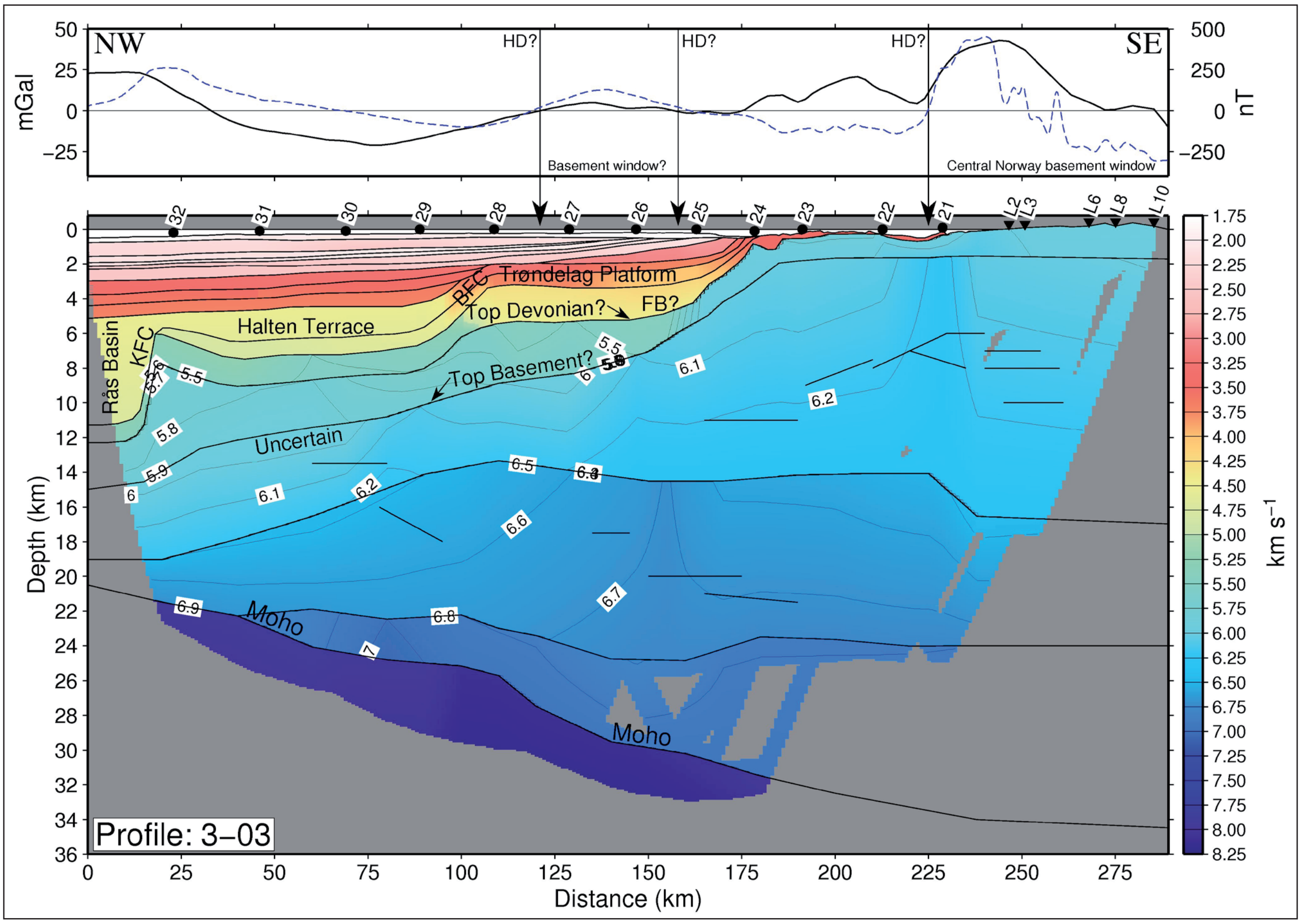


*Figure 7. Upper panel: Observed gravity- (heavy line), magnetic anomalies (dashed blue line) along Profile- 3-03. Lower panel: Gridded crustal velocity model of Profile 3-03. The parts of the model not covered by rays are masked. Floating reflectors are not included in the ray-coverage. Contour interval within the basement is 0.1 km s$^{-1}$. The intersections of the profile (Fig. 1) with the proposed offshore extension of the Høybakken Detachment (HD) (Skilbrei et al. 2002) are indicated. BFC: Bremstein Fault Complex, FB: Froan Basin, KFC: Klakk Fault Complex.*

assigned uncertainties of ±150-200 ms. We list the number of picks, $\chi^2$, and RMS ΔT fit for the arrivals constraining the sedimentary strata, the crystalline basement, and the Moho in Table 1. The $P_MP$ phases (Moho reflections) constrain both crustal velocity and Moho position, and are included in both categories.

Fit statistics indicate the ease of picking arrivals, and the ability of the model to reproduce these. Another question is how well different parts of the model are constrained, and here the ray coverage is useful. The more rays that cover a cross-section, the better the constraints on the velocity structure. We show ray hit counts for 2.5x0.25

Table 1: Fit statistics for the phases constraining main parts of the velocity models.

| Profile 3-03 | No. data points | RMS ΔT (ms) | Normalized $\chi^2$ |
|---|---|---|---|
| Sedimentary strata | 528 | 75 | 0.595 |
| Crystalline basement (incl. $P_MP$) | 3263 | 143 | 0.730 |
| Moho and upper mantle ($P_MP$ & $P_n$) | 328 | 201 | 0.995 |
| **Profile 4-03** | **No. data points** | **RMS ΔT (ms)** | **Normalized $\chi^2$** |
| Sedimentary strata | 607 | 86 | 0.862 |
| Crystalline basement (incl. $P_MP$) | 2424 | 134 | 0.929 |
| Moho and upper mantle ($P_MP$ & $P_n$) | 519 | 135 | 0.779 |

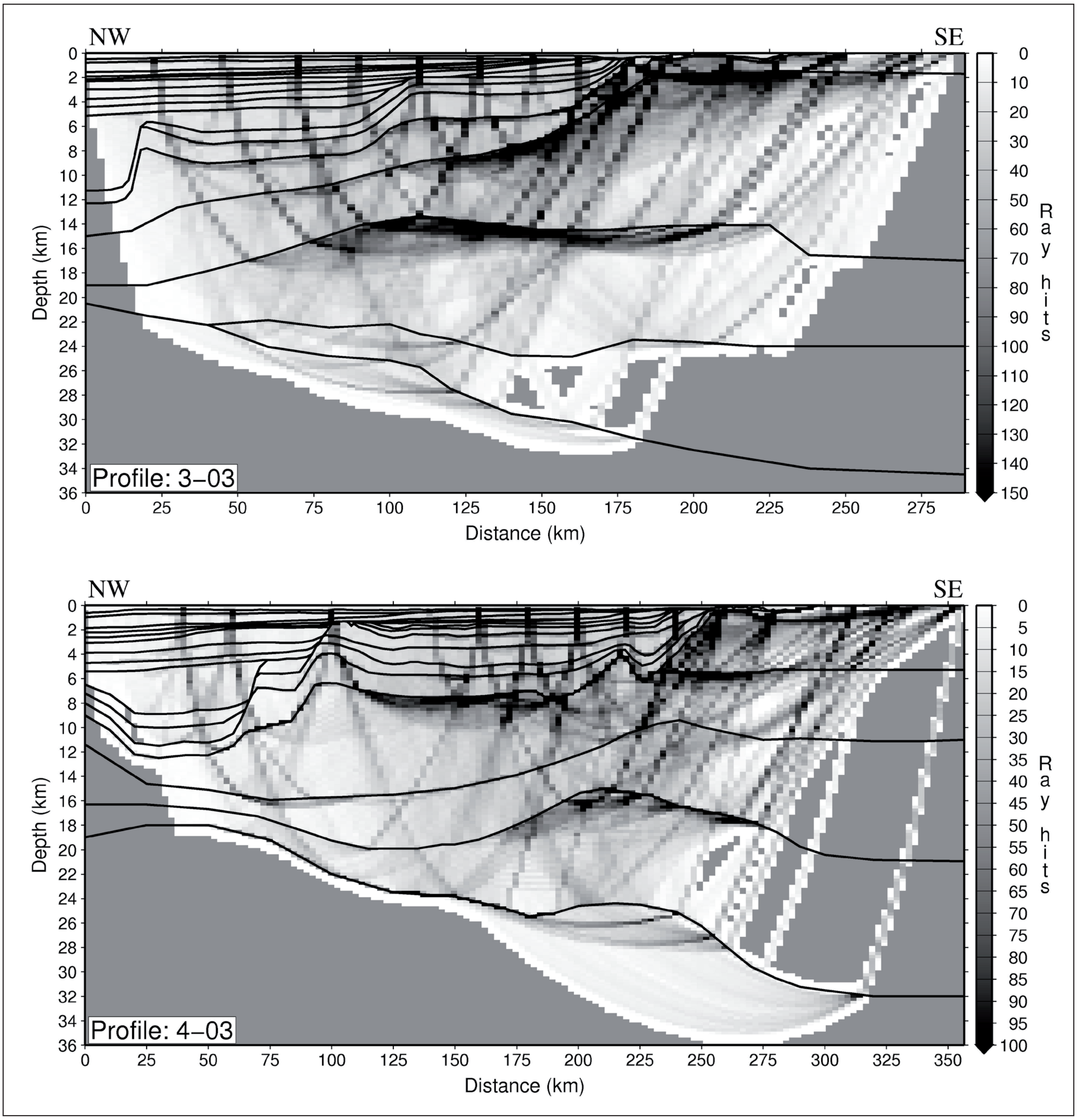


*Figure 8. Ray hit count for 2.5x0.25 km grid cell size for Profile 3-03 and 4-03. Rays from floating reflectors are not included.*

km (distance-depth) grid cell size for each model in Figure 8. On Profile 3-03 the ray coverage is dense for the upper and middle crystalline crust between 50 and 230 km, while the lower crust and Moho have best coverage between 50 and 130 km. Similarly, upper basement coverage is best between 100 and 330 km on Profile 4-03, and mid-lower crustal coverage is best between 150 and 270 km. Note the shadow zones at the ends of the models, increasing with depth.

## Gravity modeling

The gravity field is subdued at the Trøndelag Platform (Fig. 9). Central parts are slightly negative, while both northwestern ends of the profiles cover small, positive gravity anomalies. There is a 90 mGal anomaly north of the western part of Profile 4-03, above the Utgard High. There is a marked negative anomaly onshore at the southeastern end of the profile. The CNBW itself is tied to positive gravity anomalies of typically 40-50 mGal. Not all geological structures outlined by Blystad et al. (1995)

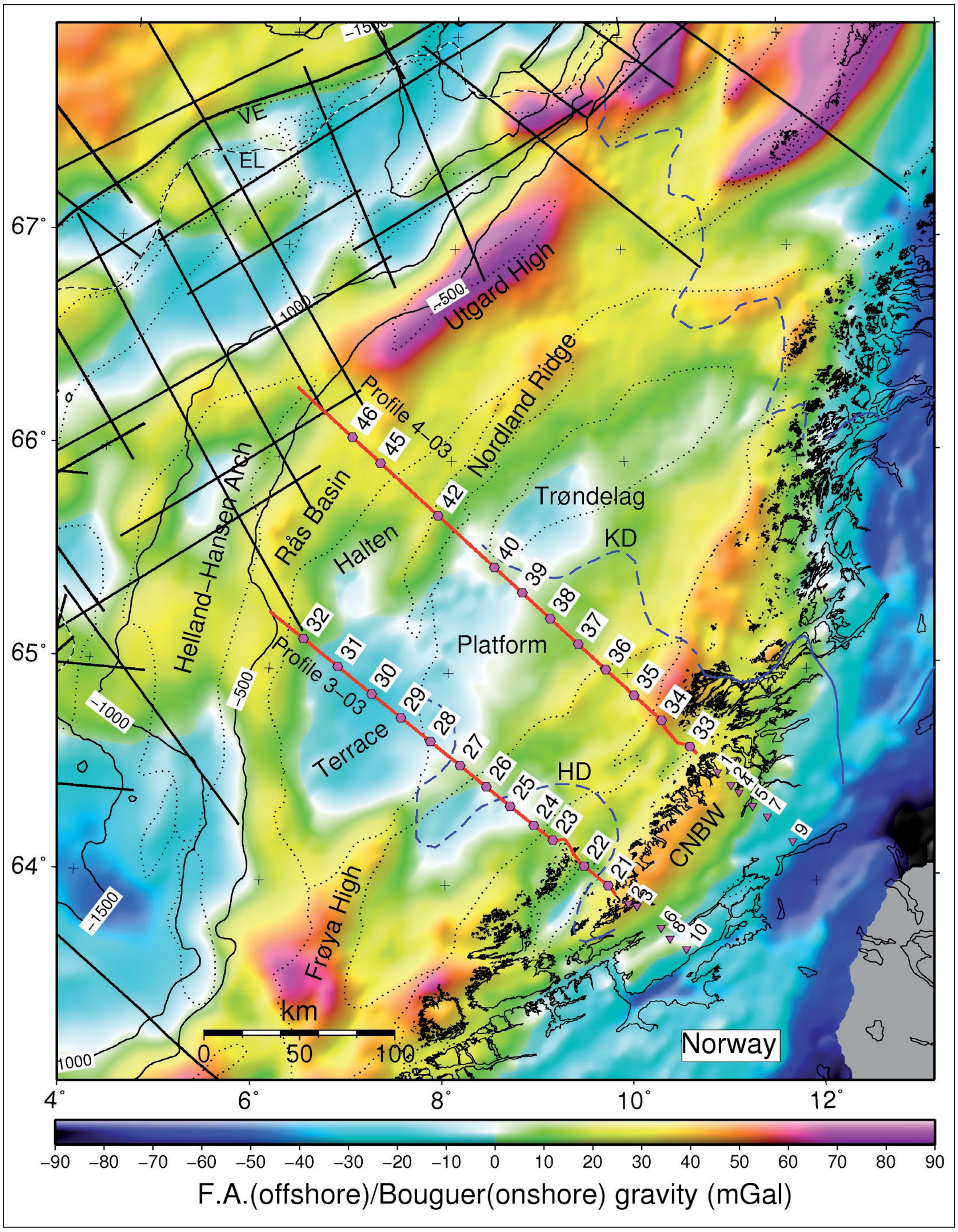


*Figure 9. Gravity map based on ERS-1 and Geosat satellite data, ship track data, and land measurements (Skilbrei et al. 2000). Offshore gravity is Free-Air, while onshore data is Bouguer corrected (using density 2670 kg m$^{-3}$). Survey navigation from the present study (red) with instrument positions are shown together with older OBS survey navigation (black). For reference, the 500, 1000, and 1500 m bathymetry contours are also shown together with proposed extensions of outcropping detachments at basement level on the shelf (Skilbrei et al. 2002; Olesen et al. 2002). Shaded relief is illuminated from the NW. CNBW: Central Norway Basement Window, EL: Eastern limit lava flows, VE: Vøring Escarpment. Outlines of structural elements (dotted) are from Blystad et al. (1995) (Fig. 1).*

show on the gravity map (Fig. 1). The Klakk FC, Revfallet FC, and the Ytreholmen FZ do not correlate to gravity gradients, and the Rås Basin does not have the expected negative gravity signature, but rather a positive one.

Initial gravity models were constructed from the velocity models using a standard, empirical relationship between velocity and density based on Ludwig et al. (1970), as summarized by Barton (1986). Since we compare to Bouguer gravity onshore, the elevation of the model inland was set to 0 km.

*Profile 4-03*

The Rås Basin has a positive gravity signature, despite being an ~11 km deep basin with a thick Cretaceous section of moderate density (Fig. 10). The seismic data (Fig. 3) indicate ~7.2 km s$^{-1}$ near top basement underneath the basin (Fig. 6). Even with the shallow Moho (~18 km), the observed gravity cannot be reproduced without high density basement. The velocity model limits the extent of the high-velocity body to the west of the Ytreholmen FZ. Also, the gravity is highest adjacent to this fault zone, indicating that the high-density body

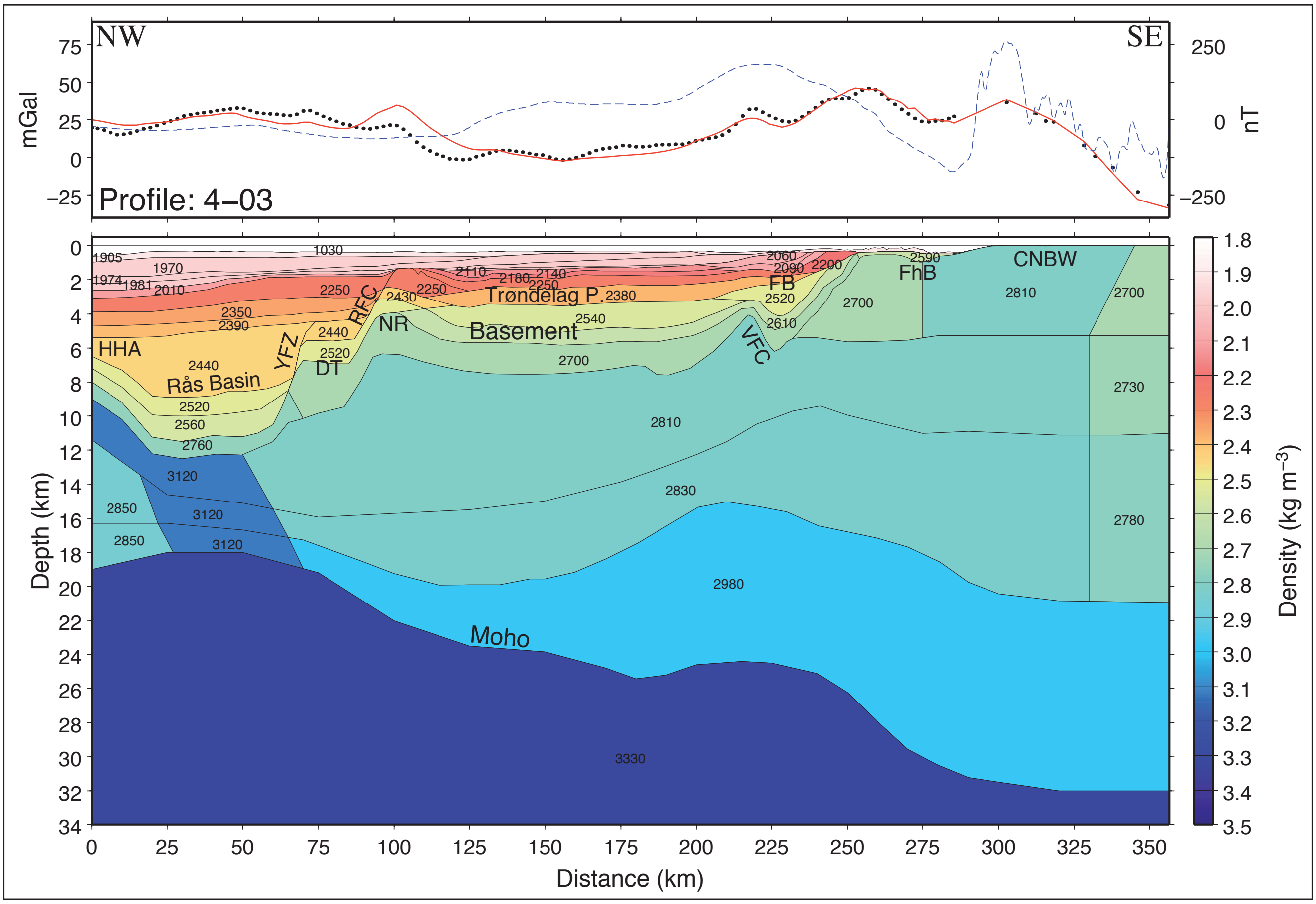


*Figure 10. Gravity model based on the P-wave velocity model of Profile 4-03 (Fig. 6). Annotated densities are in kg m$^{-3}$. Lower panel: Density distribution within the crust. Upper panel: Observed gravity (dots), calculated gravity (red line), magnetic anomalies (dashed blue line). CNBW: Central Norway Basement Window, DT: Dønna Terrace, FB: Froan Basin, FhB: Frohavet Basin, HHA: Helland-Hansen Arch, RFC: Revfallet Fault Complex, VFC: Vingleia Fault Complex, YFZ: Ytreholmen Fault Zone.*

thins to the northwest. We use a density of ~3120 kg m$^{-3}$, and let the body occupy most of the basement in the southeastern part of the Rås Basin, but only the upper part farther northwest.

The calculated gravity over the Nordland Ridge is slightly higher than that observed. Even if the velocity indicates that density could rise with stratigraphy in lower parts of the ridge, this produces a stronger gravity response than observed. The Froan Basin has a distinct, but slightly lower calculated gravity than observed. Nevertheless, the main features are reproduced, including the broad gravity- low across the Trøndelag Platform.

There is a steep fall in the observed gravity anomaly at the southeastern end of the profile. Seismic constraints on crustal thickness and basement velocity are lacking in this part (Fig. 8). Increasing crustal thickness could reproduce the low gravity, but not the steep gradient. The gravity low must correspond at least partly to a density decrease in the upper to middle crust. The observed gravity anomaly can be reproduced by introducing a body with granite density extending from the surface down to 20 km depth. Granites or equivalent volcanic rocks are exposed at the surface, supporting this solution.

*Profile 3-03*

The gravity rises from a negative anomaly over the Halten Terrace to a positive anomaly over the deeper Rås Basin (Fig. 11). The seismic data do not cover basement west of the Klakk FC, but Moho depth is constrained to 21 km nearby under the westernmost Halten Terrace. By continuing the rising Moho trend to underneath the Rås Basin, and using high basement density (3120 kg m$^{-3}$), the amplitude of the gravity is reproduced, but not the shape. Unconstrained variations in Moho depth, combined with a complex shape to the high-density body may account for this.

The transition from the Halten Terrace to the Trøndelag Platform across the Bremstein FC has no gravity signature. The local gravity high peaks ~30 km farther southeast, near the Vingleia FC. This zone was not identified in the velocity model, and therefore does not show in the

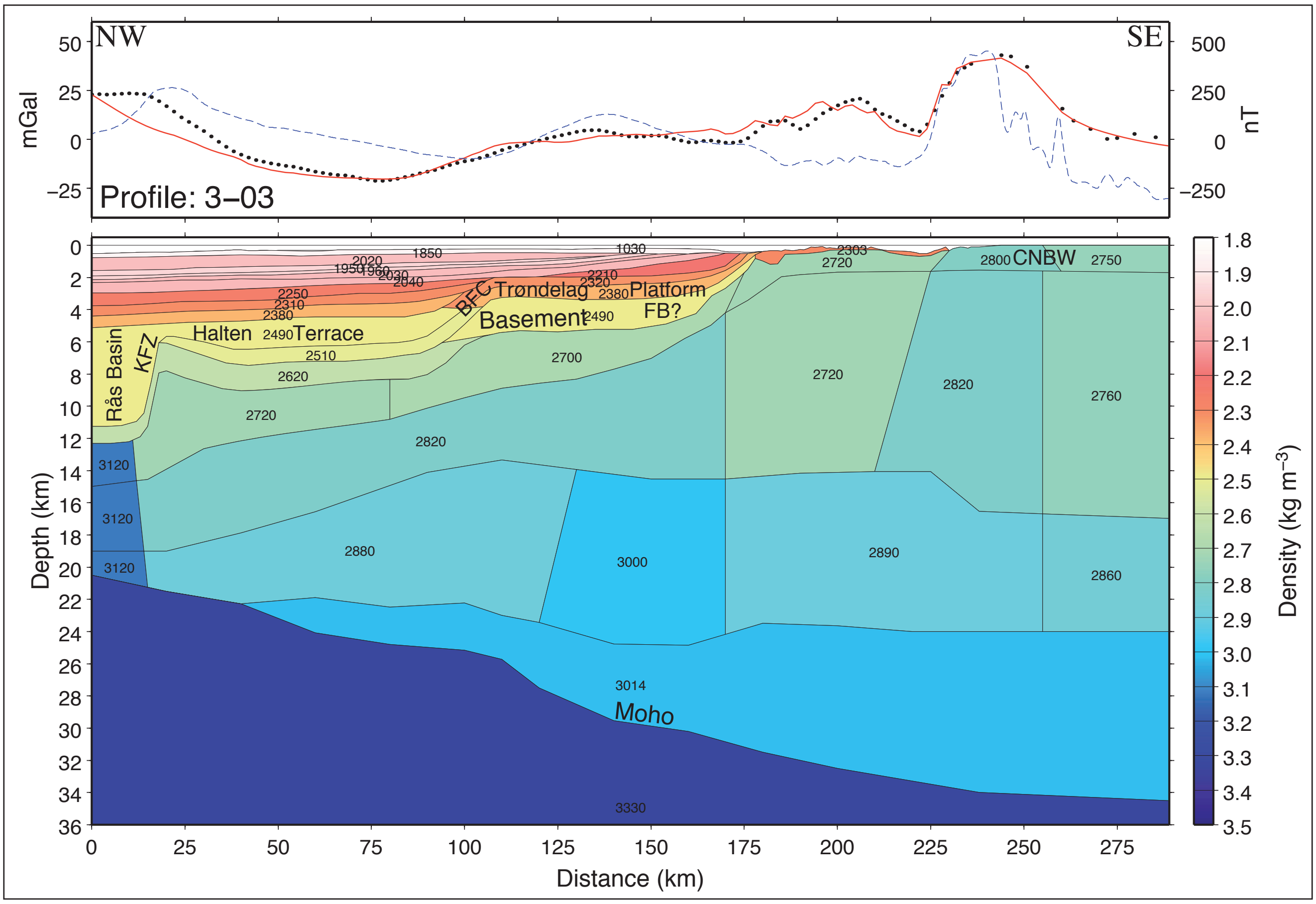


*Figure 11. Gravity model based on the P-wave velocity model of Profile 3-03 (Fig. 7). Annotated densities are in kg m⁻³. Lower panel: Density distribution within the crust. Upper panel: Observed gravity (dots), calculated gravity (red line), magnetic anomalies (dashed blue line). BFC: Bremstein Fault Complex, CNBW: Central Norway Basement Window, FB: Froan Basin, KFC: Klakk Fault Complex.*

gravity model. The edge of the Froan Basin shows as a small gravity high, even if there is no top basement structure (Fig. 9). The gravity signature may come from a minor upper basement heterogeneity.

The basement rises towards the coastline with a moderate increase in gravity, but the highest anomaly is located onshore over the CNBW. The gravity gradient is steep, starting just outside the coast. The top basement in this area shows a velocity increase from 5.90-5.95 km $s^{-1}$ to 6.05 km $s^{-1}$ under the gravity anomaly (Fig. 7). At 1.5-1.6 km depth, the increase is from 6.00-6.05 km $s^{-1}$ to 6.25 km $s^{-1}$, which is well constrained by a number of instruments. The density increase may go from 2700 to 2820 kg $m^{-3}$ here (Fig. 11). The observed gravity field falls towards the southeastern end of the profile, but not as much as on Profile 4-03 to the north. A corresponding moderate decrease in density concentrated to the upper and middle crust is consistent with this (Figs. 7 and 8).

## Discussion

### *Correlation with rocks onshore*

We show magnetic anomaly and gravity profiles both with the velocity (Figs. 6 and 7) and the density models (Figs. 10 and 11). The magnetic map over the area shows that the CNBW has a strong, positive magnetic anomaly (Fig. 12), correlating well with the positive gravity anomaly (Fig. 9). The magnetic anomalies show short wavelength fluctuations onshore southeast of the peak anomalies, obviously caused by a shallow source.

In an attempt to correlate the seismic data with exposed basement, we have prepared a map indicating three groups of basement rocks; felsic, intermediate, and mafic (Fig. 13, Tab. 2). This map is based on the outcrop map published by NGU (Sigmond 2002).

It is apparent that the strongest of the short wavelength magnetic anomalies correlates with a tongue of mafic Caledonian rocks (Cambrian-Ordovician greenstone, greenschist, amphibolite, meta-andesite) on Profile 3-03. These mafic inliers are not detectable in the

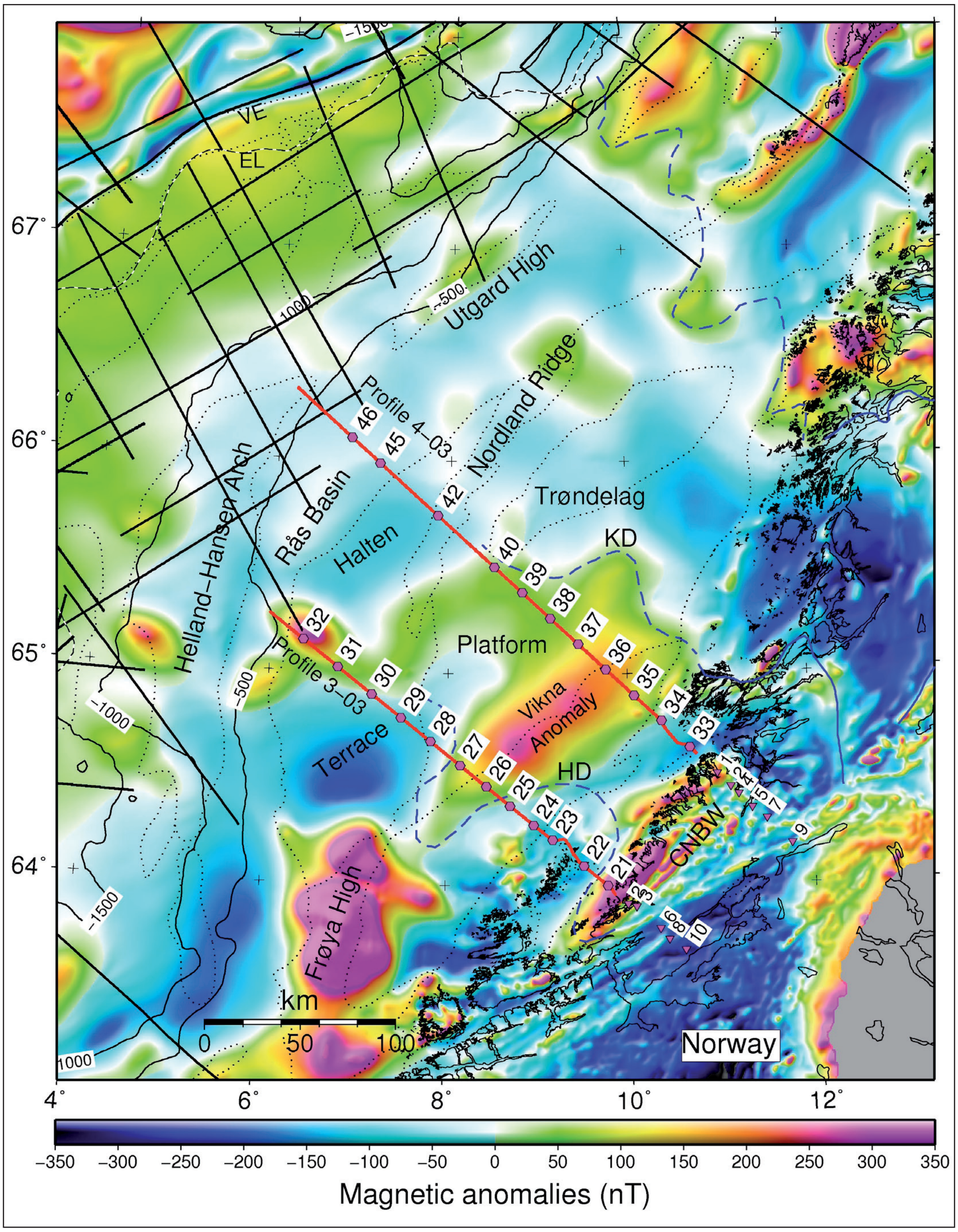


*Figure 12. Magnetic map (Olesen et al. 1997) over the study area with present (red) and older (black) OBS survey navigation. Shaded relief is illuminated from the NW. CNBW: Central Norway Basement Window, EL: Eastern limit lava flows, VE: Vøring Escarpment. Outlines of structural elements (dotted) are from Blystad et al. (1995) (Fig. 1).*

subsurface velocity distribution, and are likely to be thin. The highest upper-basement velocities on Profile 3-03 are located between the land stations and OBS 21. This area consists of mixed high-grade metamorphic Precambrian rocks, in which the magnetic anomalies are also highest.

Both gravity and magnetic anomalies fall towards the southeast on both profiles. The gravity modeling of Profile 4-03 indicates lower-density basement at the southeastern end. There is a basement window adjacent to the profile end exposing rhyolites, trachytes (volcanic equivalents of granites and syenites) and granitic gneiss of TIB (Trans-Sandinavian Igneous Belt) age. This unit is allochthonous (Sigmond 2002), though it is not necessarily transported very far. Since transport direction is to the southeast, the basement window should be representative of the bedrock underneath the end of the profile. Within the TIB, Early Proterozoic granitoids extend from southern Sweden to Lofoten (e.g., Andersen et al. 2009). Granitoids are also exposed east of the Caledonian nappes in Sweden at this latitude. Olesen et al. (2002) concluded from gravity modeling that a massive granite complex underlies the nappes in northern Nordland, locally reaching ~20 km depth, similar to what our model indicates for mid-Norway.

Both gravity and magnetic anomalies fall in the near

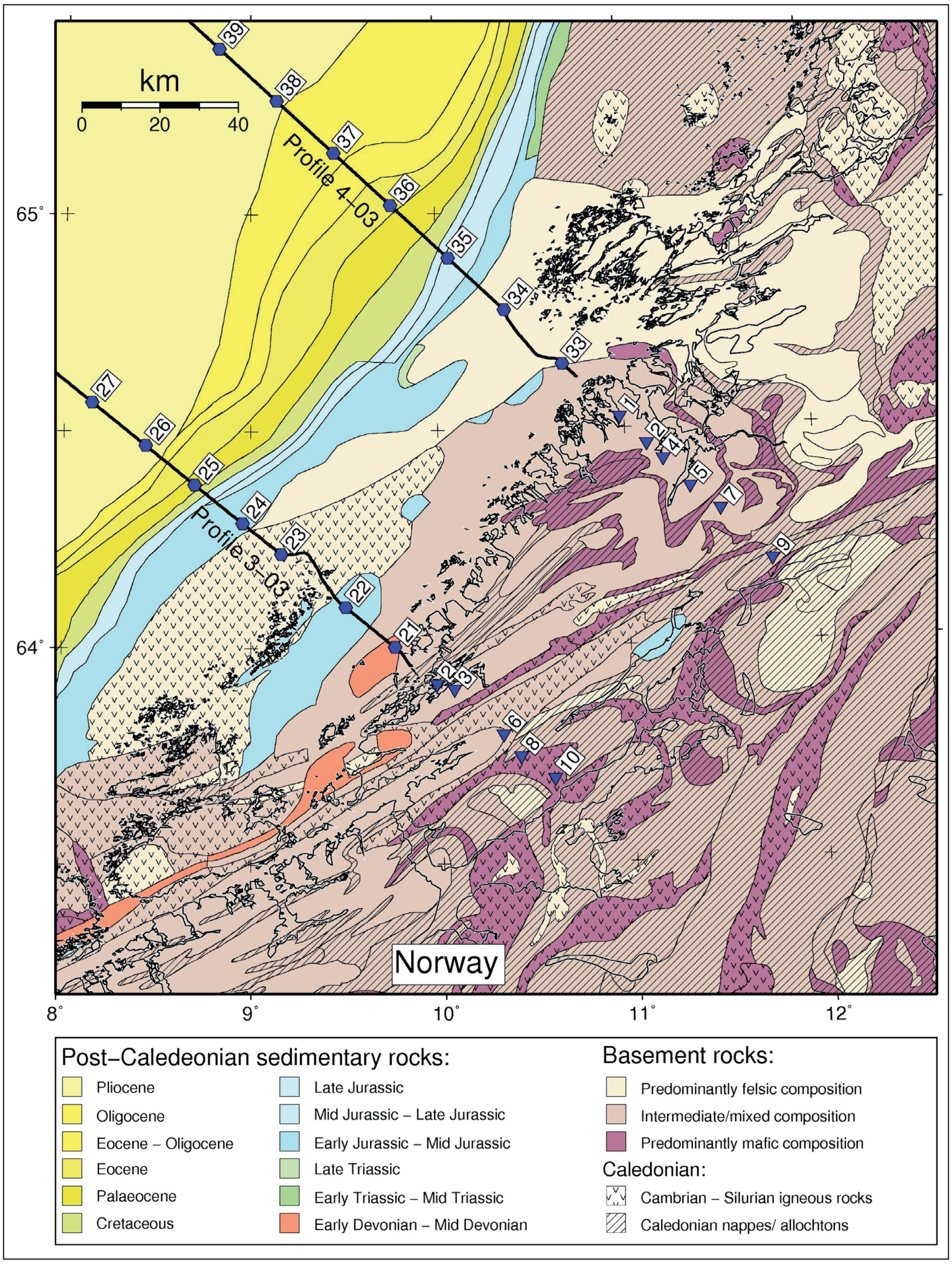


*Figure 13. Rock outcrop map of near shore and onshore areas near the seismic land stations. Onshore rocks have been grouped into felsic, intermediate and mafic rocks, for comparison with seismic properties (Tab. 2). The map is based on the digital distribution of the NGU rock outcrop map (Sigmond, 2002).*

offshore areas, indicating a reduction in upper basement density and magnetization. The outcrop map indicates predominantly Caledonian felsic intrusive rocks (Sigmond 2002), consistent with this. Farther offshore the magnetic field in particular rises again, as is discussed below.

*Offshore continuation of late- to post-Caledonian detachments?*

Magnetization of the crystalline basement rocks is closely tied to the metamorphic grade. The strongest magnetization correlates with granulite facies rocks of intermediate to mafic composition, as seen in the Lofoten Archipelago (Schlinger 1985). This is due to the amphibolite facies reactions which do not facilitate the growth of magnetite. The granulite facies rocks of the CNBW (e.g., Braathen et al. 2002) are consistent with this tied to a strong, positive magnetic anomaly, and the detachments at the contact with Caledonian nappes to the north and south border this anomaly. It was therefore assumed that the magnetic anomaly pattern could be used to extend the detachments out onto the shelf: The offshore extension of the Kollstraumen Detachment (Olesen et al. 2002) and the Høybakken Detachment (Skilbrei et al. 2002) were traced at each side of a positive magnetic anomaly northwest of the CNBW (Fig. 12). There are, however, some problematic as well

Table 2: List of the composition of various rock units as published by Sigmond (2002), loosely grouped by dominant quartz content as shown in Fig. 13.

| |
|---|
| **Felsic rock units** |
| Rhyolite, trachyte, leptite, granitic gneiss |
| Granite, in part porphyritic, syenite, quartz syenite, monzonite, augen granite |
| Granitic gneiss, foliated granite, locally paragneiss |
| Augen gneiss, foliated porphyritic granite |
| Rhyolite, trachyte |
| Gneiss, migmatite, granite, augen gneiss, mica gneiss, metamorphosed supracrustal rocks |
| Granite, granodiorite, norite |
| Metarhyolite, metarhyodacite, quartz keratophyre |
| **Intermediate rock units** |
| Mica gneiss, mica schist, phyllite, quartzite, amphibolite, marble, migmatitic gneiss |
| Gneiss, migmatite, amphibolite, foliated granite, mica gneiss |
| Mica schist, mica gneiss |
| Meta- sandstone, conglomerate, limestone, shale, volcanics, phyllite, mica schist, quartzite, tillite |
| Quartz sandstone, siltstone, claystone, limestone, dolostone, conglomerate, tillite |
| Phyllite, mica schist, mica gneiss, metasandstone, amphibolite, conglomerate |
| Limestone, marble |
| Quartz sandstone, greywacke, phyllite, metasiltstone, limestone, dolostone, marble |
| Basic to acidic volcanic rocks |
| Trondhjemite, tonalite, granodiorite |
| Diorite, quartz diorite |
| **Mafic rock units** |
| Gabbro, diorite, amphibolite, anorthosite and mangerite |
| Amphibolite, locally greenschist, metagabbro, diabase, mica schist |
| Gabbro, diorite, quartz diorite, monzonite, monzodiorite, peridotite |
| Greenstone, greenschist, amphibolite, meta-andesite |

as implicit assumptions connected to this interpretation. To represent detachment faults by a trace on a map, the fault plane must be truncated by another plane, presumably an erosional surface. Onshore, the contact is easily seen, but offshore sedimentation has buried Caledonian structures, and the detachments as drawn must outcrop at top basement under the sedimentary succession. The choice of magnetic contour level to correlate with the presumed outcrop is not discussed, but appears close to 0 nT. Buried by several km of sedimentary deposits, with all other parameters being equal the level should be different from the level onshore. The magnetic interpretation of the detachment trace is not very accurate even near shore, where it is placed west of Asenøya, while field data restrict it to lie between the island and the mainland (Eide et al. 2005, Osmundsen et al. 2006). Also, magnetization within the CNBW is highly variable, and does not support the view that there should be a consistent magnetic level to follow (Fig. 12). Furthermore, there is a poor continuation from the onshore to the offshore magnetic anomalies, suggesting that the magnetic source is not continuous between the areas.

Profile 4-03 lies entirely within the proposed offshore basement window, while Profile 3-03 crosses the proposed Høybakken Detachment extension three times (Fig. 12). If there is a sharp boundary between amphibolite and granulite facies rocks across a detachment, we expect this to be detectable in the seismic velocity (Hurich et al. 2001). We also note a good correlation between increased seismic velocity, magnetic and gravity anomalies onshore within the CNBW along Profile 3-03 (Fig. 7), though it is less clear for Profile 4-03 (Fig. 6). The first crossing of the Høybakken Detachment by Profile 3-03 is between OBSs 21 and 22 (HD, Fig. 7). The top basement velocity is ~5.9 km $s^{-1}$ and comparable to that of the top basement of the CNBW to the southeast, but there is no velocity contrast. The next crossing occurs between OBSs 25 and 26, where top basement again should be Precambrian to the northwest. In this position, there is a rapid fall in velocity down to 5.4-5.5 km $s^{-1}$, indicating the first appearance of the well-consolidated sedimentary rocks. This transition could represent a Devonian detachment with Devonian sedimentary rocks to the northwest. Top basement is then under this layer. Basement velocities here do not stand out from surrounding areas, except by being slightly lower. The last crossing occurs between OBSs 27 and 28, where there are no upper basement changes in velocity. Thus, the seismic data do not confirm the predicted offshore continuation of the detachments.

Profile 4-03 shows the same top basement trends as Profile 3-03 (Fig. 6). The top basement velocity is highest under land stations 1 to 7 (6.10-6.15 km $s^{-1}$), and the maximum magnetic amplitude is located at the northwestern edge of this zone. Top basement velocity decreases to ~6.0 km $s^{-1}$ on the inner part of the shelf. Between OBSs 34 and 35, a layer of lower velocity (~5.5 km $s^{-1}$) appears. By OBS 36, at the outer edge of the Froan Basin, the layer is ~2 km thick. The thickness may increase towards the Revfallet FC, but this is less clear than the thickness increase seen on Profile 3-03. It is likely that this layer, at least in part, consists of Devonian strata as discussed above. Profile 4-03 should be located entirely within the buried basement window. The broad peak of the offshore magnetic anomaly used to define the detachments does not correlate with any top or upper basement seismic structure. On the other hand, it does correlate with a distinct lower-crustal body. We will discuss this in the next section.

It appears that the CNBW was denuded by bidirectional movement on the Høybakken and Kollstraumen detachments, with the Møre-Trøndelag FC acting as a transfer zone (Braathen et al. 2000, 2002). Denudation was substantial and apparently triggered some magmatism, producing pegmatite dikes. On the other hand, Precambrian basement is exposed southeast of the MTFC, indicating that deep erosion also played a role. While the magnitude of movement is difficult to assess, extensional faulting is likely to have affected areas now offshore. However, we find no evidence that detachments are exposed in the top of the basement offshore. This does not preclude these detachment faults from continuing offshore, but using the magnetic field to interpret detachment traces out on the shelf is not reliable for a number of the reasons pointed out above. On the other hand, our data indicate a near offshore detachment accommodating Devonian strata towards the northwest.

While there are no convincing upper basement structures that correlate with the offshore magnetic high, the lower crust is quite distinct under the Trøndelag Platform. The top has a marked velocity increase of 0.3-0.5 km $s^{-1}$ from the crust above, at a depth of 14-16 km. On Profile 4-03, the top is dome-shaped, and located directly under the peak anomaly (Fig. 6). Here, beneath the Froan Basin, we also find the highest velocities (6.70-6.85 km $s^{-1}$) and the greatest thickness (8-9.5 km) of the lower crust. Profile 3-03 has the highest velocities beneath the inner Halten Terrace (Fig. 7). Velocities in the upper part of the lower crust are slightly lower here (6.60-6.75 km $s^{-1}$), but the thickness (~15 km) is greater, and the basal crustal velocities are higher (6.90-7.05 km $s^{-1}$). The highest velocity is offset slightly to the southeast from the peak magnetic anomaly, and the top of the lower crust is flatter. Based on a magnetic model, Ebbing et al. (2009) proposed that basement with high magnetization outcrops beneath the Froan Basin, but our seismic models indicate that the source exists lower in the crust.

The velocity of the lower crust indicates granulite facies, intermediate to mafic rocks (e.g., Holbrook et al. 1992; Hurich et al. 2001). This is consistent with high magnetization and the increased density in the gravity models. The origin is not easily determined, however: It could represent an island arc accreted during an earlier orogeny. If it dates to the Svecofennian orogeny, it could be an igneous intrusion of intermediate to mafic composition, as seen in connection with the TIB farther south (Korja & Heikkinen 2005; Korja et al. 2006). The TIB granites were created through melting of the crust caused by heat input at its base, requiring massive igneous intrusion into the lower crust (Andersen et al. 2009). The location adjacent to a massive granite intrusion is typical for similar lower-crustal bodies seen on the BABEL lines in the Baltic (Korja & Heikkinen 2005). The seismic velocities of the lower crust do not indicate elevated mantle temperatures during melting: Formation is likely related to the Early Proterozoic subduction believed to cause the formation of the TIB province.

### *Early Eocene igneous underplating?*

Large amounts of intrusive and extrusive igneous rocks were produced over a few million years in connection with continental breakup at the mid-Norwegian Møre and Vøring margins (e.g., Eldholm & Grue 1994; Berndt et al. 2001; Svensen et al. 2004; Mjelde et al. 2005b; Breivik et al. 2006, 2009). High velocity (7.1 to 7.7 km $s^{-1}$) lower crust has been identified over most of the outer Vøring Margin, locally reaching more than 8 km in thickness (Mjelde et al. 2005a, 2009). Mantle melting under elevated temperatures will produce magnesium-rich gabbros that can explain velocities in the 7.1-7.3 km $s^{-1}$ range (White et al. 2008). Higher velocities require other rock types. However, the velocity in itself does not give a unique identification of the rocks present, which could also be Caledonian (or older) mafic high-grade metamorphic rocks (Gernigon et al. 2004, 2006; Ebbing et al. 2006). This difference is fundamental for the interpretation of on the tectonic and thermal evolution of the sedimentary basins offshore, and subsequently the petroleum potential.

The only identification of similar velocities (~7.2 km $s^{-1}$) in our data is in the Rås Basin on Profile 4-03 (Fig. 3). The gravity model indicates high density throughout the crust (Fig. 10), and the gravity map shows a strong, positive gravity anomaly at the adjacent Utgard High (Fig. 9). Two older OBS lines indicate high lower crustal velocity here (Mjelde et al. 1998), and it is likely that the anomalous basement is continuous with the rocks coring the high. The gravity anomaly continues in an arc southwards to the Frøya High. Profile 3-03 ends on this anomaly, where the gravity model similarly indicates high basement density. It could therefore be a continuous, older mafic terrane which was thinned during formation of the Rås Basin, but is still thick in the Utgard and Frøya Highs. We note, however, that the magnetic signature of the Utgard High is moderate, while it is strong at the Frøya High, and that the continuation of the magnetic anomaly is poor between the two highs (Fig. 12).

The ties between our profiles and older OBS surveys have little overlap (Fig. 1). As explained earlier, ties between OBS profiles need several tens of kms overlap in order to be good in the lower crust. The crustal control beneath the Rås Basin is therefore much poorer than the line coverage suggests. Our results indicate shallower Moho depth (18-21 km) than the tie profiles (~25 km) (Raum et al. 2002). However, we trust our result which is consist-ent with crustal thinning from the platform to the basin. Profile 4-03 ties to the older 1992 OBS survey (Mjelde et al. 1997), which indicated a ~5 km thick LCB within the tie area (Mjelde et al. 2005b, 2009), assumed to be an Early Eocene igneous underplating/intrusion. However, we now believe it is more likely to be mafic Caledonian or older rocks tied to the Utgard High. This interpretation does not of course preclude an Eocene igneous LCB farther west, though we find no clear evidence of any igneous underplating of the Rås Basin, the Halten Terrace, or the Trøndelag Platform.

*Caledonian suture?*

While the Caledonian suture and thrust zones have been recognized elsewhere (e.g., Brewer & Smythe 1984; Klemperer & Hobbs 1991; Gudlaugsson & Faleide 1994; Breivik et al. 2005), a suture has not been found on the Norwegian mainland or in the shelf areas. On conventional multichannel seismic data, major thrust zones and sutures can be seen as changes in crustal reflectivity and/or thickness over the zone. These zones are often associated with reflections continuing into the upper mantle-, and may correlate with a crustal root (e.g., Klemperer & Hobbs 1991). In the absence of later tectonism, such structures can be inherited even from the Proterozoic (e.g., Korja & Heikkinen 2005). Wide-angle seismic data may show the crustal root, sometimes with increased upper mantle velocity below, as seen south of Svalbard (Breivik et al. 2005), or in the Urals (e.g., Carbonell et al. 2000). No signs of a suture have been identified on the Norwegian shelf.

Our data show two basement bodies that could represent accreted exotic terranes within a continent-continent collision zone. If the inner intermediate-to-mafic basement block is an island arc caught up in the Caledonian orogeny, the Laurentian terrane should border this. The other mafic body is located farther west, possibly forming a continuous arc from the Utgard High, under the Rås Basin, to the Frøya High. This could be an island arc or oceanic terrane, and if so it was more likely to have been accreted during the Caledonian orogeny than the inner block. The suture could then be located at or west of this arc. The extreme extension of the Vøring Basins (e.g., Brekke 2000; Mjelde et al. 2005b) may have overprinted any structures from the Caledonian terrane accretion, making them difficult to detect.

*Comparison to seismic interpretation*

It is not always straight forward to compare velocity models with MCS data, as velocity is primarily a function of the compaction and diagenetic history, and not the stratigraphy. At the Profile 4-03 location the Rås Basin appears more asymmetric on MCS data, with the deepest part towards the arch (Figs. 2B and 6). Strata dip strongly from the Helland-Hansen Arch into the basin, down to ~4-5 km greater depth than indicated by the velocity transect. These strata lie deep and are probably well consolidated, and may not easily be recognized in the velocity model. The deepest part could be pre-Cretaceous strata, which would leave more space for older deposits, as the crust is thin (~18 km) here.

Strata on the inner part of the Dønna Terrace dip towards the Nordland Ridge, while strata on the outer part dip towards the Rås Basin (Fig. 2B). Correlation across the terrace is uncertain, but the velocity model indicates that the largest fault throw could be located at the edge of the inner fault block. This makes the Ytreholmen FZ wider, and the Dønna Terrace 15-20 km narrower in this position than indicated in Fig. 2B.

On the inner part of the shelf, the Froan Basin is distinct in the north at Profile 4-03, but has no basement signature at Profile 3-03 in the central part of the basin. The northern and southern parts of the basin (Fig.1) may be separate structures, and the Vingleia FC may not be a continuous fault system from south to north.

## Summary and conclusions

Two wide-angle seismic profiles were acquired from the Rås Basin to the onshore Central Norway Basement Window (CNBW) under the Euromargins program in 2003. The data were collected by Ocean Bottom Seismometers and by land-based seismic stations. The starting model was derived from older MCS data, providing reliable stratigraphic interpretation down to top Triassic. Crustal velocity models that can reproduce the observed seismic travel times were constructed using forward/inverse ray-tracing software (Zelt & Smith 1992). As there is a good correlation between rock velocity and density, forward gravity modeling was used to constrain crustal structure in parts poorly covered by seismic arrivals.

Top basement velocity at the CNBW is 5.95-6.15 km $s^{-1}$. At ~70 km from the coast on Profile 3-03, there is an abrupt change from 5.95 to 5.4 km $s^{-1}$ in the upper ~1.5 km of the basement. This is interpreted as a Devonian detachment, and the intermediate-velocity layer likely consists mostly of well-consolidated Devonian deposits. The thickness increases towards the Ytreholmen FZ to the northwest, and Late Paleozoic-Mesozoic rifting clearly post-dates this layer. The depth to Devonian strata is then 5-6 km beneath the Trøndelag Platform and 7-8 km beneath the Halten Terrace, but the depth to crystalline basement is partly obscured by this layer. Basement depth is estimated to be 7-9 km at the Trøndelag Platform, but uncertain under the Halten Terrace. Basement depth in the Rås Basin is only constrained by Profile 4-03, where it is 11-12 km.

Moho depth is ~22-25.5 km under the central Trøndelag Platform, decreasing to 18-21 km under the Rås Basin. Moho depth increases from ~24 km at the Halten Terrace to 26-30 km at the southern Trøndelag Platform, while it is 31-32 km on both profiles near the coastline.

There is a distinct lower crustal body under the Trøndelag Platform, being thickest under the Froan Basin. The top of the body lies at a depth of 12-14 km and is a strong reflector, consistent with the marked increase in velocity from overlying basement rocks. Characteristic velocities lie in the 6.6 to 7.0 km $s^{-1}$ range, indicating granulite facies rocks of intermediate-to-mafic composition. The shape and dimension resemble Proterozoic mafic nuclei or intermediate-to-mafic intrusions connected with the development of the Trans-Scandinavian Igneous Belt (e.g., Korja & Heikkinen 2005).

High seismic velocity (~7.2 km $s^{-1}$) is found in the upper part of the basement under the Rås Basin. Gravity modeling indicates that this is a high-density body occupying most of the basement, and that it is also present farther south, adjacent to the Klakk FC. This body causes the gravity field to increase from the platform to the Rås Basin, despite the depth increase. The high density is located near the Utgard High to the north, which has a strong, positive gravity anomaly. Possible continuity with this basement suggests that the crust under the Rås Basin is not related to the igneous underplating farther west. Earlier interpretations suggested ~5 km underplating south of the Utgard High (Mjelde et al. 2005b, 2009). Our results, seen in a regional context, indicate that this could rather be an island arc or oceanic terrane accreted by the Caledonian orogeny, and possibly continuous from the Utgard High to the Frøya High. There are no indications of typical LCB under the platform areas.

The Høybakken and Kollstraumen detachments record orogen-parallel extension, and are thought to have unroofed the CNBW during collapse of the Caledonides (e.g., Braathen et al. 2000, 2002). As the CNBW has quite strong magnetization, it has been proposed that the edges of a distinct, positive magnetic anomaly (Vikna anomaly) offshore show the seaward continuation of these detachments (Olesen et al. 2002; Skilbrei et al. 2002). Top basement should then be high-grade Precambrian basement, also under the platform area. This is not confirmed in our profiles, which show moderate upper basement velocities. At our southern profile, the transition from typical basement velocity to lower velocity occurs at a crossing of the proposed extension of the Høybakken Detachment, but with opposite polarity to the expected. This is likely a Devonian detachment, accommodating thick Devonian strata to the northwest on the shelf. The Vikna anomaly appears to correlate with the distinct lower crustal block rather than with the upper basement structure, and is therefore not suitable for mapping offshore detachments.

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